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# Issues concerning atmospheric turbidity indices

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ABSTRACT

Knowledge of atmospheric turbidity coefficients is very important in meteorology, climatology, atmospheric pollution monitoring, and in the prediction of solar energy availability under cloudless skies. This paper provides a thorough review on a number of atmospheric turbidity indices and on the several methods that have been developed in the past few decades. The Lambert–Bouguer–Beer law is the basic relationship underlying the derivation of various turbidity indices. Turbidity can be referred to as monochromatic (narrow band) wavelengths, broadband wavelengths, and the total spectrum. Narrow band turbidity is measured using sun photometers, while the other two turbidities are measured using pyrheliometers with broadband pass filters. The Ångström's turbidity coefficients and the Linke's turbidity factor are among the most frequently used atmospheric turbidity coefficients.


**Contents**



## 1. Introduction

Most energy produced in the fusion furnace of the sun is radially transmitted as electromagnetic radiation, which is popularly called sunshine or solar energy [1]. The sun has a temperature of 5760 K, and it emits the total power of approximately $3.8 \times 10^{30}$ W or $6.25 \times 10^{11}$ W m$^{-2}$. Since the distance between the sun and the earth is large, the amount of solar radiation that reaches the outside of the earth's atmosphere is quite low at only 1367 W m$^{-2}$, which is called the total solar irradiance (TSI) or the solar constant [2]. The solar constant is measured on the surface perpendicular to the sun's rays at the average sun–earth distance above the atmosphere [3]. The intensity of solar radiation above the atmosphere approximately varies sinusoidally over the year with amplitude close to 3.3% of the solar constant and it achieves its maximum around the first day of January. This variation arises from the variation in the distance of the earth from the sun [4]. Variation in solar radiation is also caused by the difference in the emission intensity from the sun itself [3].

Not all solar radiation received at the periphery of the atmosphere reaches the Earth's surface [5]. Solar radiation is partially depleted and attenuated as it traverses the atmospheric layers [6].

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**Nomenclature**

| | |
|---|---|
| $\beta$ | the Ångström's turbidity coefficient |
| $\alpha$ | the Ångström wavelength exponent |
| $\tau_{\alpha\lambda}$ | spectral aerosol optical depth at the wavelength $\lambda$ |
| $Vis$ | visibility in kilometers |
| $I_d$ | diffuse solar radiation (W m$^{-2}$) |
| $I_b$ | direct solar radiation (W m$^{-2}$) |
| $I_0$ | the irradiance of the extraterrestrial solar radiation |
| $I_{Sc}$ | extraterrestrial solar irradiance at normal incidence (1367 W m$^{-2}$) |
| $I_{cs}$ | threshold irradiance |
| $I_{bn}$ | normal direct solar radiation at the earth's surface (W m$^{-2}$) |
| $E_0$ | correction term for the earth–sun distance |
| $I^*_{bn}$ | normal incidence solar radiation at the bottom of a dust-free atmosphere |
| $\tau_5$ | the aerosol optical thickness at 0.5 μm |
| $\beta_G$ | Ångström turbidity coefficient obtained using Gueymard method (dimensionless). |
| $m_r$ | relative optical air mass at a given pressure (dimensionless) |
| $T_{Pin}$ | Linke turbidity factor obtained using Pinazo et al. method (dimensionless) |
| $K_{db}$ | the ratio between the diffuse and direct beam normal irradiance |
| $\lambda_{aff}$ | the effective wavelength corresponding to each of the filters |
| $\tau_{a\Delta\lambda}$ | the aerosol optical thickness at a specified spectral band |
| $T_L$ | Linke's turbidity factor (dimensionless) |
| $T_{Lm}$ | monthly average values of Linke's turbidity factor |
| $\delta_R(m_r)$ | the integral Rayleigh optical thickness |
| $\delta_R(z)$ | the relative optical thickness relating to Rayleigh scattering by the gaseous molecules |
| $\delta_D(z)$ | the relative optical thickness associated with aerosol extinction and gaseous absorption other than ozone |
| $\delta_{RK}(m_r)$ | the integral Rayleigh optical thickness computed by the formula of Kasten [63] |
| $T_{LK}(m_r)$ | Linke's turbidity factor determined by Kasten's algorithm [63] |
| $T_{La}(m_r)$ | Linke's turbidity factor determined by Kasten's algorithm [68] |
| $\delta_{Ra}(m_r)$ | integral Rayleigh optical thickness computed by the formula of [68] |
| $T_{Lvis}$ | Linke's turbidity factor in the visible region of the solar spectrum |
| $I_{hn}$ | direct hourly normal solar radiation |
| $T_{LIP}$ | Linke turbidity factor obtained using Pinazo et al. method (dimensionless) |
| $\overline{\tau}_R$ | the mean Rayleigh extinction |
| $\overline{\tau}_w$ | the selective absorption coefficient for atmospheric gases |
| $T_U$ | Unsworth–Monteith turbidity coefficient |
| $T_{LAM2}$ | Linke's turbidity factor for an air mass equal to 2 |
| $w$ | precipitable water thickness (cm). |
| $K_{db}$ | the standardised value (to zero altitude and total column ozone equal to 0.3434 atm cm) of the ratio between diffuse and direct irradiance |
| $B$ | Schüepp's turbidity coefficient |
| $m^*$ | an absolute air mass |
| CDA | the number of clean-dry atmospheres |
| $\gamma$ | solar elevation angle |
| $T_{il}$ | illuminance's turbidity factor |
| $T_V$ | Volz's turbidity factor |
| $\tau^R_\lambda$ | Rayleigh scattering coefficient for air moleculares at the wavelength $\lambda$ |
| $\tau^{oz}_\lambda$ | absorption coefficient for ozone at $\lambda$ |
| $T_K$ | Kastrov's turbidity coefficient |
| $T_r$ | rational's turbidity factor |

The presence of aerosol particles in the atmosphere is one of the most important factors that affect the amount of solar radiation that reaches the Earth's surface under cloudless skies [7]. Aerosols are small particles that are either in the solid or liquid state [8]. Common aerosol particles have radius ranging from 1 to $10^5$ nm, while very small particles (called Aitken particles) range from 1 to 10 nm and large particles range from $10^2$ to $10^3$ nm. Particles in the range of $10^3$ to $10^5$ nm are called giant particles [9,10]. Based on the formation of aerosol dispersions, aerosol particles have two categories, namely, dispersed and condensed aerosols [8]. Aerosols have two dominant layers in the atmosphere. The first dominant layer is near the Earth's surface at 0–3 km, which is affected by natural dust storms and man-made inputs to the atmosphere, whereas the second dominant layer is the stratospheric dust layer at 15–25 km above sea level, which is caused by volcanic action and cosmic sources [11].

Aerosols have direct radiative forcing because they scatter and absorb solar and infrared radiations in the atmosphere. Aerosols also alter warm, ice, and mixed-phase cloud formation processes by increasing droplet concentrations and ice particle concentrations, consequently causing indirect radiative forcing associated with the changes in the microphysical and optical properties of clouds. Furthermore, aerosols are highly absorbent of solar radiation, such as black carbon, and they may reduce cloud cover and liquid water contents by heating the cloud and the environment where the cloud forms. This is known as the semi-direct effect because it is the result of the direct interaction of aerosols with radiation that indirectly influences the climate by altering clouds. The net effect of aerosols on climate depends on the summation of all the three mechanisms above [12]. An atmosphere that contains aerosols is described as turbid or hazy. An aerosol-laden atmosphere has the property called atmospheric turbidity, which depletes direct solar radiation [10].

Studies on atmospheric turbidity are important in meteorology, climatology, and in atmospheric pollution monitoring. Atmospheric turbidity parameters are necessary for determining the amount of spectral global irradiance and for designing photovoltaic cells and selective absorbers for spectral thermal collectors [13]. Aerosols make the atmosphere turbid, and thus, the presence of aerosols is commonly quantified by an index of turbidity. In the last century, some indices of turbidity were proposed and several techniques were developed to determine their values [14]. Among these indices are the Ångström's and the Schüepp's turbidity coefficients, which are spectral in nature, and the Unsworth–Monteith's coefficient and Link's turbidity factor, which are broadband in nature [15]. This study mainly aims to investigate atmospheric turbidity and then describe available techniques for calculating various turbidity indices.

## 2. Atmospheric aerosol characteristics

The word aerosol was introduced more than 80 years ago as an analogy to the term hydrosol, a stable liquid suspension of solid

particles. Aerosols are products of a complicated totality of chemical and physical processes [16,17]. Aerosols are defined in their simplest forms as small solid or liquid particles that remain suspended in the air and follow the motion of the air within certain broad limits. An atmosphere that contains aerosols is described as turbid or hazy. An aerosol-laden atmosphere has the property called atmospheric turbidity, which depletes direct solar radiation [16,10]. The presence of aerosols in the Earth's atmosphere has important effects on the transmission of solar radiation and on the radiative heat transfers in the atmosphere. This is due to the fact that aerosols absorb and scatter solar radiation as it passes through the atmosphere [14]. Scattering refers to processes in which photons change direction after an interaction, while absorption occurs when photons are removed from a beam light and their energy is converted to an excitation of atoms or molecules [18]. Aerosols are physically divided into the following three categories according to particle size:

1. Aitken nucleus with particle radius smaller than $10^2$ nm
2. Large nucleus with particle radius larger than $10^2$ nm and smaller than $10^3$ nm
3. Giant nucleus with particle radius larger than $10^3$ nm

Based on the formation of the aerosol dispersions, aerosols fall into the following two categories [8]:

1. *Dispersed aerosols*: the aerosol materials that are either in the solid or liquid state are pulverized into a granular state via mechanical pulverization or natural weathering, and are then suspended in the air because of the increase in wind force.
2. *Condensed aerosols*: the formation of the aerosol particle or droplet is caused by the condensation of super-saturated gas on the condensation nucleus or by the mixture of different gases through photochemical reaction.

Studies on the physical properties of aerosol particles are important for the assessment of the effects aerosol particles on the climate and for the development of more accurate remote-sensing procedures from satellite sensors [19]. Aerosol properties vary with time and from one region to another because of their short lifespan and highly temporal and spatial variability measurements [20,21]. The spatial-temporal variability of aerosol characteristics is large and the observational data are so fragmentary that the total content of various types of aerosols is difficult to estimate, while the available power estimates of the global sources of natural and anthropogenic aerosols are rather approximate [17].

Kondratyev et al. [17] and Giannakaki et al. [22] reported that the types of aerosols depend on their sources and are roughly categorized into anthropogenic or natural. Anthropogenic sources are determined by human activities, such as industrial wastes from chimneys, toxic exhausts from cars, fires, explosions, soil erosion in agriculture, and open mining. The estimates of anthropogenic aerosols are more reliable compared with those of natural aerosols, especially in regions along the World Ocean and in the continents that are difficult to access. Atmospheric aerosols are classified into the following types of natural aerosols based on their composition or sources:

1. Products of sea spray evaporation.
2. Mineral dust wind-driven to the atmosphere.
3. Volcanic aerosols (both directly emitted to the atmosphere and formed by gas-to-particle conversion).
4. Particles of biogenic origin (directly emitted to the atmosphere and formed by the condensation of volatile organic compounds, such as terpenes, and from the chemical reactions between these compounds).
5. Smokes from biota burning on land.
6. Products of natural gas-to-particle conversion (e.g., sulfates from reduced sulfur on the ocean surface with emissions of dimethyl sulfide).

Important types of anthropogenic aerosols also include the following:

1. Direct industrial emissions of particles (soot, smoke, and road dust, among others).
2. Products of gas-to-particle conversion.

The amount of aerosols in the atmosphere is sometimes measured in terms of the number of particles per cubic centimeter. The number of dust particles is usually greater on land than over water and during the drier seasons. Moreover, the number of dust particles is lower in cold polar or arctic air and is higher in tropical air. The number density exponentially decreases with altitudes of up to approximately 5 km and remains constant at altitudes of 10–15 km. The number density slightly increases in a layer centered around 20 km because of the thermal and dynamic nature of the atmosphere [10]. Tropospheric aerosol particles play an important role in the attenuation of solar radiation because they strongly affect the Earth's radiation budget. The influence of aerosols on radiation that passes through the atmosphere cannot be neglected, especially in urban or industrialized areas, [20,21,23].

Aerosols are a minor constituent of the Earth's atmosphere, but they play an important role in the energy balance of the earth–atmosphere system. Aerosols can influence the climate in two ways [24]. The magnitude of these influences, however, is poorly constrained because of limited knowledge on the processes that control the distributions as well as the physical, chemical, and optical properties of aerosols [25]. First, aerosol particles can reduce solar energy by absorbing or scattering the radiation under cloud-free and cloudy conditions, thereby affecting the energy balance of the Earth [24]. This direct effect is related to the optical properties of aerosol particles, such as single scattering albedo, size distribution, size variability with relative humidity, complex refractive index, and solubility. The levels of scattering and absorption depend on the physical and chemical characteristics of aerosol particles. Consequently, aerosols modify the Earth's radiation budget and consequently influence the warming or cooling of the planet. Second, aerosols alter the formation and the precipitation efficiency of liquid-water, ice, and mixed-phase clouds, thereby causing indirect radiative forcing associated with these changes in the cloud properties. The quantification of indirect radiative forcing caused by aerosols is an especially difficult task [26].

Radiative forcing of the climate by aerosols may be comparable in magnitude, but opposite in sign, to that of greenhouse gases. Greenhouse gases trap radiation and warm the surface of the earth, while aerosols absorb and scatter radiation back to space, theoretically resulting in a net cooling effect [24]. This variability is largely due to the much shorter atmospheric lifetime of aerosols compared with that of greenhouse gases [23]. Turbidity observations had been made since 1923, but their measurements are inherently difficult and expensive even making some of them not possible because of the strong influence of other atmospheric components [27,15,28]. The accurate determination of turbidity normally requires clear-sky spectral radiation data obtained using sun photometers or spectroradiometers. These instruments are expensive and scarce, and thus, turbidity is

instead generally estimated from broadband irradiance measurements [29].

## 3. Atmospheric turbidity indices

Aerosols have a relationship with the attenuation of solar radiation that reaches the Earth's surface, and thus, a number of methods and atmospheric turbidity indices have been introduced and developed in the past decades to evaluate atmospheric turbidity and to quantify its influence on the direct irradiance on the Earth's surface [14,30,31]. The Ångström turbidity coefficients $\alpha$ and $\beta$ and the Schüepp turbidity coefficient $B$, among others, are spectral in nature, such that they reference the spectral aerosol optical depth to a common wavelength (i.e., 1000 nm for $\beta$ and 500 nm for $B$). On the other hand, the Unsworth–Monteith turbidity coefficient $T_U$ and the Link turbidity factor $T_L$ are broadband in nature, i.e., they are estimated by reversing or inverting the models that estimate broadband beam irradiance upon measurement. The Unsworth–Monteith turbidity coefficient is, by definition, equal to the broadband aerosol optical depth, which is the sum of aerosol absorption and scattering coefficients integrated over the entire depth of the atmosphere [15,30]. The definition of each turbidity index is explained in the succeeding sections.

### 3.1. Ångström's turbidity coefficients ($\alpha$ and $\beta$)

Under clear sky conditions, atmospheric aerosols generally contribute to the largest attenuation of solar radiation in the visible range. Scattering and absorption effects cause the attenuation. Since attenuation effects of the scattering and absorption by aerosols are difficult to separate, Ångström suggested a single formula that is generally known as the Ångström turbidity formula, which can be expressed as follows [10,21,32]:

$$\tau_{\alpha\lambda} = \beta\lambda^{-\alpha} \tag{1}$$

where $\tau_{\alpha\lambda}$ is the spectral aerosol optical depth at the wavelength, $\beta$ is the Ångström turbidity coefficient, $\alpha$ is the wavelength exponent, and $\lambda$ is the wavelength in micrometer.

In Eq. (1), $\beta$, which varies from 0.0 to 0.5 or even higher, is an index that represents the amount of aerosols in the atmosphere at the vertical direction. The wavelength exponent $\alpha$ is related to the size distribution of the aerosol particles. Large values of $\alpha$ indicate a relatively high ratio of small to large particles, which vary from 0 to 4. When the aerosol particles are very small in relation to air molecules, $\alpha$ should approach 4, whereas for very large particles, $\alpha$ should approach 0. Generally, $\alpha$ is has a value between 0.5 and 2.5. The average $\alpha$ for the entire solar spectrum has a value between 0.9 and 2.0. An acceptable average value for most natural atmospheres is $\alpha = 1.3 \pm 0.5$. Table 1 lists the combinations of parameters $\beta$ and $\alpha$, which may be considered for various degrees of atmospheric cleanliness [10,33].

The Ångström exponent $\alpha$ is a widely used parameter in atmospheric sciences that examines the optical properties of aerosol particles. Since the early [34,35] and later [36,37] publications of Ångström, where the $\alpha$ parameter was mainly used to describe the spectral behavior of atmospheric extinction and transmission, $\alpha$ is now also applied to a variety of similar but slightly different optical properties such as the atmosphere scattering or backscattering coefficients. The Ångström exponent is very popular because of the simplicity of its equation that enables the interpolation or extrapolation of aerosol optical properties and because of its connection to particle microphysics (related to the mean size of aerosols) [38].

**Table 1**
Parameters for various degrees of atmospheric cleanliness [10].

| Atmosphere | $\beta$ | $\alpha$ | Visibility (km) |
|---|---|---|---|
| Clean | 0.00 | 1.30 | 340 |
| Clear | 0.10 | 1.30 | 28 |
| Turbid | 0.20 | 1.30 | 11 |
| Very turbid | 0.40 | 1.30 | < 5 |

#### 3.1.1. Techniques for determining the Ångström turbidity coefficients ($\alpha$ and $\beta$)

Ångström turbidity coefficients $\alpha$ and $\beta$ can be determined using a number of techniques. Traditionally, $\alpha$ and $\beta$ are derived from either spectral direct solar radiation measurements or from broadband direct solar radiation and precipitable water measurements [39,40].

*3.1.1.1. The Volz method.* Volz [41] applies Eq. (1) to sun photometric measurements performed in two different narrow bands that are chosen to minimize the influence of other atmospheric attenuators, such as water vapor. In this case, $\alpha$ can be obtained as follows:

$$\alpha = \frac{d\ln\tau_{\alpha\lambda}}{d\ln\lambda} = -\frac{\ln\left(\frac{\tau_{\alpha_2}}{\tau_{\alpha_1}}\right)}{\ln\frac{\lambda_2}{\lambda_1}} \tag{2}$$

where $\lambda_1$ and $\lambda_2$ represent the effective wavelengths in the corresponding narrow bands [21,26].

The chosen wavelengths are usually 380 and 500 nm, where weather conditions play a very important role in the distribution of pollutants in the air [39,28]. At 380 nm, no molecular absorption is observed, while at 500 nm, the ozone has weak absorption [10]. Eq. (2) indicates that $\alpha$ is the negative of the slope or the negative of the first derivative of $\tau_{\alpha\lambda}$ versus the wavelength in logarithmic scale. After $\alpha$ is derived, $\beta$ can be calculated in the same wavelength interval using the values of $\alpha$ and $\tau_{\alpha\lambda}$ at a certain $\lambda$ [21]. However, $\beta$ alone can be measured with a single-wavelength Volz instrument by assuming that $\alpha = 1.3$. Alternately, $\beta$ can be measured at $\lambda = 1$ μm, where the exponent $\alpha$ is not included in the calculations. At this wavelength, only the water vapor has a weak absorption band. This method is very accurate and yields the values of $\alpha$ and $\beta$ simultaneously [10,39,28].

*3.1.1.2. Direct method.* The direct method enables the calculation of the Ångström turbidity coefficients $\alpha$ and $\beta$ by using the following formula obtained from Eq. (1) [42]:

$$\alpha = -\frac{d[\ln\tau_{\alpha\lambda}]}{d\ln\lambda} = -\frac{\lambda}{\tau_{\alpha\lambda}}\left[\frac{d\tau_{\alpha\lambda}}{d\lambda}\right] \tag{3}$$

The parameter $\beta$ can also be determined from Eq. (1).

*3.1.1.3. Linear fitting.* Linear fitting depends on the omission of $\tau_{\alpha\lambda}$ in the strong absorption bands of water vapor and mixed gases, and it determines the Ångström parameters at remaining wavelengths. This fact constitutes the main difference with the previous (direct) method. The Ångström parameters are derived by applying the least-squares fit, as written in Eq. (1), to the remaining wavelengths, which can be rewritten in the following linear form [42]:

$$\ln\tau_{\alpha\lambda} = -\alpha\ln\lambda + \ln\beta \tag{4}$$

In this approximation, the slope of the straight line yields $\alpha$, while the intercept provides $\beta$. Hence, $\alpha$ is the negative of the slope (first derivative) of $\ln\tau_{\alpha\lambda}$ versus $\ln\lambda$. This method has often been used for determining aerosol characteristics because of its

simplicity [21,43]. Since the Ångström formula does not exactly fit the data points, a second-order polynomial fit for ln $\tau_{\alpha\lambda}$ versus ln $\lambda$ data provides excellent agreement with the measured $\tau_{\alpha\lambda}$, resulting in order differences of the uncertainty of the measurements (~0.01–0.02), while a linear fit yields significant differences with the measured $\tau_{\alpha\lambda}$. The second-order polynomial fit is applied to the $\tau_{\alpha\lambda}$ values at all wavelengths and is expressed in the following form [26]:

$$\ln\tau_{\alpha\lambda} = \alpha_2 \ln\lambda^2 + \alpha_1 \ln\lambda + \alpha_0 \tag{5}$$

Since $\alpha$ is the negative of the first derivative of ln $\tau_{\alpha\lambda}$ versus ln $\lambda$, the following expression can be derived from Eq. (4) to quantify the spectral dependence of $\alpha$:

$$\alpha = -2\alpha_2 \ln\lambda - \alpha_1 \tag{6}$$

*3.1.1.4. Horizontal visibility.* Routine measurements of $\beta$ and $\alpha$ with a dual-wavelength sun photometer are conducted at a number of locations. When such measurements are unavailable, the turbidity parameter $\beta$ may be determined from the measurement of visibility (Vis) in the horizontal direction, if such a measurement exists. This visibility is also called the meteorological range. For visibilities greater than 5 km, the turbidity parameter can be determined using the following equation developed by McClatchey, Selby, and others [10]:

$$\beta = 0.55^{\alpha}(3.912/Vis - 0.01162)[0.02472(Vis - 5) + 1.132] \tag{7}$$

where *Vis* is in kilometers. In order to predict $\beta$, the value of $\alpha$ must first be known. Unfortunately, even with the help of proper instruments, the measured visibility is slightly subjective in nature. Therefore, the relationship above only yields a gross estimate of $\beta$.

*3.1.1.5. Rangarajan and Mani method.* Rangarajan and Mani [44] described a new, simpler, and less expensive method of evaluating $\beta$ from direct and diffused solar radiation measurements. Their method has been a routine in principal radiation stations.

$$\frac{I_d}{I_b} = \frac{0.5S_a + 0.75S_d}{(1-S_a)(1-S_d)} \tag{8}$$

$$S_d = 1 - g(\beta)^{m_r} \tag{9}$$

$$S_a = \frac{0.606m^*}{643 + m^*} \tag{10}$$

$$g(\beta) = 1 - 1.375\beta \quad \text{for} \quad \beta < 0.08 \tag{11}$$

$$g(\beta) = 0.9843 - 1.1429\beta \quad \text{for} \quad \beta \geq 0.08 \tag{12}$$

where $m^*$ is an absolute air mass, $I_b$ is the direct solar radiation, and $I_d$ is the diffused solar radiation.

*3.1.1.6. Louche et al. method.* Louche et al. [39] described a method for determining the Ångström's turbidity coefficient in a Mediterranean location from the total pyrheliometric measurements in the parameterization of the direct solar total irradiance by Bird and Hulstrom [45,46], which can be expressed as follows:

$$\beta = \frac{1}{m_r D}\ln\left(\frac{C}{A-B}\right) \tag{13}$$

where $m_r$ is the relative air mass at actual pressure (dimensionless) and

$$A = I_{bn}/(0.975E_0 I_{Sc}\tau_r\tau_o\tau_g\tau_w)$$

$$B = 0.1244\alpha - 0.0162$$

$$C = 1.003 - 0.125\alpha$$

$$D = 1.089\alpha + 0.5123$$

where $E_0$ is a correction term for the earth–sun distance.

*3.1.1.7. Gueymard model.* Gueymard [47,48] developed a new model for solar beam spectral irradiance. This model uses the following two possible $\alpha$ values: $\alpha_1$ for $\lambda < 0.5$ μm and $\alpha_2$ for other measurements. The $\beta$ parameter is independent of the wavelength and is derived from the following relation:

$$\beta_G = 0.5^{\alpha_2}|\tau_5 \tag{14}$$

where $\tau_5$ is the aerosol optical thickness at 0.5 μm. Consequently, only a single measurement of the beam spectral irradiance at this wavelength is necessary.

*3.1.1.8. Pinazo et al. method.* Pinazo et al. [40] presented a new method (referred to as Method II) for the calculation of the Ångström turbidity coefficient based on the ratio of direct solar radiation to global solar radiation on a horizontal surface and on the "C" model of Iqbal. Method II permits the calculation of $\beta$ without employing certain variables, such as the thicknesses of the precipitable water and of the ozone layer. Furthermore, measurements on dry temperature and relative humidity are not necessary. According to Method II, $\beta$ can be expressed as follows:

$$\beta_{Pin} = -\ln\frac{(D - 0.12445\alpha + 0.0162)/(1.003 - 0.125\alpha)}{m_r(1.089\alpha + 0.5123)} \tag{15}$$

The definition of *D* can be found in the original paper of the author [40].

*3.1.1.9. Gueymard and Vignola method.* Gueymard and Vignola [29] proposed a semi-physical method to evaluate turbidity from broadband irradiance measurements and from other atmospheric parameters. This method should be under cloudless skies and requires very accurate measurements of global (or diffused) and direct radiations. From the number of parametric runs performed with SMARTS2, $K_{db}$ ($K_{db}=I_d/I_{bn}$; the ratio between diffused and direct beam normal irradiance) can be accurately parameterized as follows:

$$K_{db} = \left(a_0 + a_1\beta + a_2\beta^2\right)/\left(1 + a_3\beta^2\right) \tag{16}$$

where the coefficients $a_i$ are functions of the zenith angle, pressure ozone amount, and precipitable water.

If an experimental value of $K_{db}=I_d/I_{bn}$ is obtained from the measurements, Eq. (16) can easily be solved to obtain $\beta$ as follows:

$$\beta = 0.5\left\{\left[a_1^2 - 4(a_2 - a_3K_{db})(a_0 - K_{db})\right]^{0.5} - a_1\right\}/(a_2 - a_3K_{db}) \tag{17}$$

*3.1.1.10. Malik method.* Malik [7] modified the Pinazo et al. model [40] to introduce a new variable, called the "horizontal visibility," to more accurately estimate the turbidity coefficient of Ångström $\beta$ from the ratio of horizontal direct solar irradiance to the global horizontal irradiance and horizontal visibility.

$$\beta = -\ln\frac{(D - 0.12445\alpha + 0.9162)/(1.003 - 0.125\alpha)}{m_r(1.089\alpha + 0.5123)} \tag{18}$$

*3.1.1.11. Utrillas et al. method.* Utrillas et al. [19] developed a new method aimed at determining the Ångström turbidity coefficients

by measuring broadband direct irradiance by using filters.

$$\ln\tau_{\alpha\Delta\lambda} = \ln\beta - \alpha\,\ln\lambda_{aff} \tag{19}$$

where $\tau_{\alpha\Delta\lambda}$ is the aerosol optical thickness at a specified spectral band and $\lambda_{aff}$ is the effective wavelength that corresponds to each filter.

### 3.2. Linke's turbidity factor ($T_L$)

Linke [58], [59] introduced the Linke turbidity factor ($T_L$) and showed the number of dry atmospheres necessary to cause the attenuation of extra-terrestrial radiation produced by the real atmosphere [49]. $T_L$ refers to the entire spectrum or to the overall spectrally integrated attenuation [50]. Gueymard [51] suggested that the $T_L$ coefficient is not a pure turbidity coefficient because it also incorporates water vapor and $NO_2$ optical depths. Nevertheless, the $T_L$ coefficient may be useful when no data on precipitable water are available, making the determination of other coefficients difficult ([52]reported). Although $T_L$ does not explicitly show the effects of aerosols and water vapor, its easy measurement made it quite popular in meteorological stations [53].

$T_L$ is a key input for several models that assess the downwelling irradiance under clear skies. These models are used by several communities in the fields of renewable energies, climatology, agro-meteorology, and atmospheric pollution [54]. $T_L$ is also useful for the prediction of the availability of solar radiation and daylight illuminance under cloudless skies [55]. Although $T_L$ is a useful parameter in many applications, it has a serious drawback. $T_L$ varies with air mass even when atmospheric conditions remain constant [39]. $T_L$ normally varies from 1 to 10 [31]. In a pure Rayleigh atmosphere, $T_L$=1. The value closest to this ideal value is achieved in extremely clear and cold air at high latitudes, $T_L$=2. In a polluted atmosphere, $T_L$ can increase to 8 [56]. The attenuation of the radiation under a clear sky atmosphere becomes larger with increasing $T_L$.

#### 3.2.1. Techniques for determining the Linke's turbidity factor ($T_L$)

$T_L$ can be directly obtained from the pyrheliometric measurements of the beam irradiance during periods of very clear sky with an air mass of 2. Since this kind of experimental data is rarely available, $T_L$ generally becomes an estimated parameter. In addition, the time series of the radiation data is generally too short to allow estimation on a daily basis. Fortunately, long-term monthly average values $T_{Lm}$ are sufficient for most applications. Several methods for estimating $T_{Lm}$ values can be found in literature [27,57].

*3.2.1.1. Linke.* Linke [58] defined the integral Rayleigh optical thickness $\delta_R(m_r)$ as the integrated optical thickness of the terrestrial atmosphere that is free of clouds, water vapor, and aerosols. He calculated $\delta_R(m_r)$ from theoretical assumptions and validated it in a very pure and dry mountain atmosphere by using the following formulation:

$$\delta_R(m_r) = 0.128 - 0.054\,\log m_r \tag{20}$$

*3.2.1.2. Feussner and Dubois.* Feussner and Dubois [60] published a series of spectral data tables that enables the calculation of $\delta_R(m_r)$, where both molecular scattering and the absorption by the stratospheric ozone layer are taken into account.

*3.2.1.3. Valko.* Valko [61] suggested empirical fits of $T_L$ as a function of aerosol, water vapor content, and relative air mass. Valko's formula valid for $5^\circ < \gamma < 65^\circ$can be expressed as follows:

$$T_L = (B + 0.54)\left[1.75\,\log(w/m_r + 0.1) + 14.5\right] - 5.4 \tag{21}$$

where $B = 1.07\beta$ (by Schüepp).

*3.2.1.4. Dogniaux.* Dogniaux [62] derived the following relation from extensive experimental campaigns, illustrating the observed variation in $T_L$ with solar height $\gamma$ (°), atmospheric water vapor content $w$ (cm), and the Ångström's turbidity coefficient $\beta$:

$$T_L = \left(\frac{85+\gamma}{39.5e^{-w} + 47.4} + 0.1\right) + (16 + 0.22w)\beta \text{ for } 5^\circ < \gamma < 65^\circ \tag{22}$$

*3.2.1.5. Jaenicke and Kasten.* Jaenicke and Kasten [63] developed the following linear relationship to determine the atmospheric turbidity factor from the threshold solar elevation that is measured by any given sunshine autograph with known threshold irradiance $I_{cs}$:

$$T_L = 0.154\,\ln(I_0/I_{cs})\alpha + 1.05\,\ln(I_0/I_{cs}) \text{ for } \gamma\ 30^\circ \tag{23}$$

where $I_0$ is the irradiance of the extra-terrestrial solar radiation and $I_0 = 1375$ W m$^{-2}$ is the mean sun–earth distance.

*3.2.1.6. Kasten.* To facilitate the determination of $T_L(m_r)$ from pyrheliometric measurements, Kasten [64] presented the following simple analytic formula for $\delta_R(m_r)$, known as Kasten's pyrheliometric formula, based on numerical values published by Feussner and Dubois [59] and on international recommendations by Special Committee for the International Geophysical Year (CSAGI 1958):

$$\delta_R(m_r) = (9.4 + 0.9m_r)^{-1} \tag{24}$$

*3.2.1.7. Katz et al.* Katz et al. [65] experimentally discovered that $T_L = 2.0 + 19\beta$, but they used the definition by Kasten [64] on the integrated optical thickness of the terrestrial atmosphere, which only includes Rayleigh scattering and ozone absorption.

*3.2.1.8. Kasten.* $T_L$ at height $z$ meters above sea level $T_L(z)$ was objectively defined by Kasten [66] as ([61] reported) follows:

$$T_L(z) = 1 + \delta_D(z)/\delta_R(z) \tag{25}$$

where $\delta_R(z)$ is the relative optical thickness associated with Rayleigh scattering by the gaseous molecules in the atmosphere and ozone absorption, and $\delta_D(z)$ is the relative optical thickness associated with aerosol extinction and gaseous absorption other than the ozone in the stratosphere.

*3.2.1.9. Louche et al.* Louche et al. [50] used more recent data on the spectral extra-terrestrial solar radiation and on the spectral absorption coefficients of the atmospheric gases, as compiled by Iqbal [67], to calculate the improved numerical values of the integral Rayleigh optical thickness $\delta_R(m_r)$ according to the integration procedure outlined by Kasten [64]. Based on the new values of $\delta_R(m_r)$, Louche et al. proposed a new analytic formula for the function $\delta_R(m_r)$ and demonstrated that the new $\delta_R(m_r)$ values are definitely higher than the old CSAGI values used by Kasten, particularly, at small relative optical air masses $m_r$. Louche et al. also proposed an approximate equation for converting old turbidity factors $T_L(m_r)$ into new ones.

$$\delta_R(m_r) = \left(6.5567 + 1.7513m_r - 0.1202m_r^2 + 0.0065m_r^3 - 0.00013m_r^4\right)^{-1} \tag{26}$$

*3.2.1.10. Abdelrahman et al.* Abdelrahman et al. [53] conducted turbidity measurements in Dhahran, Saudi Arabia which has a

semi-arid climate. The inter-relationships between $T_L$ and $\beta$ were found to be linear, similar with that in Avignon (France) and in Potsdam (West Germany). However, the regression coefficients were different for Dhahran.

$$\beta = -0.1326 + 0.0433 T_L \quad (27)$$

*3.2.1.11. Grenier et al.* Grenier et al. [68] proposed a fourth-order polynomial for the function $\delta_R(m_r)$ that is similar with Louche et al. [50] but with slightly different coefficients. They confined the range of validity to $1 \le m_r \le 6$ to achieve a better fit in this range.

$$\delta_R(m_r) = \left(5.4729 + 3.0312 m_r - 0.6329 m_r^2 + 0.0910 m_r^3 - 0.00512 m_r^4\right)^{-1} \quad (28)$$

Furthermore, Grenier et al. illustrated the dependence of $T_L$ with air mass and proposed a Linke turbidity index that was normalized at an air mass of 2, which they called $T_{LAM2}$, and was therefore independent of air mass. They also investigated the dependence of $T_{LAM2}$ on water vapor and on the Ångström spectral turbidity factor and presented these dependencies in the form of various tables and polynomial fits.

$$\beta = -0.120 + 0.071 T_{LAM2} \quad 0.5 < w < 1.6 \text{ cm} \quad (29)$$

$$\beta = -0.128 + 0.070 T_{LAM2} \quad 1.2 < w < 3.2 \text{ cm} \quad (30)$$

$$\beta = -0.138 + 0.069 T_{LAM2} \quad w > 2.8 \text{ cm} \quad (31)$$

*3.2.1.12. Molineaux et al.* Molineaux et al. [39] noted that the expressions by Louche and Grenier for $\delta_R(m_r)$ became divergent for air mass measurements that were greater than 20 and 7. They adopted the coefficients of the original expression by Linke to take into account the absorption of permanent gases as follows:

$$\delta_R(m_r) = 0.124 - 0.0656 \log m_r \quad (32)$$

*3.2.1.13. Kasten.* To compare and judge the two parameterization formulae for $\delta_R(m_r)$ proposed by Louche et al. [50] and by Grenier et al. [68], respectively, Kasten [69] presented a new computation of $\delta_R(m_r)$ according to his outlined integration procedure described in [64], which was repeated by Louche et al. This new computation disregarded several simplifications and assumptions made by Louche et al. and by Grenier et al., while additional parameters were considered and some recently improved input data were used.

$$T_{La}(m_r) = T_{LK}(m_r)\left[1/\delta_{Ra}(m_r)\right]/\left[1/\delta_{RK}(m_r)\right] \quad (33)$$

where $T_{La}(m_r)$ is the new Linke turbidity factor, $T_{LK}(m_r)$ is the Linke turbidity factor determined by the algorithm of Kasten [64], $\delta_{RK}(m_r)$ is the integral Rayleigh optical thickness calculated using the formula of Kasten [64], and $\delta_{Ra}(m_r)$ is the integral Rayleigh optical thickness calculated using the formula of the new spectral integration, which is given by:

$$\delta_{Ra}(m_r) = \left(6.6296 + 1.7513 m_r - 0.1202 m_r^2 + 0.0065 m_r^3 - 0.00013 m_r^4\right)^{-1} \quad (34)$$

*3.2.1.14. Kambezidis et al.* Kambezidis et al. [70] derived an expression for the Linke turbidity parameter in the visible region of the solar spectrum $T_{Lvis}$. Their expression is based on newly identified visible transmittances for various atmospheric constituents obtained by spectrally integrating wavelength-dependent atmospheric transmittances.

$$T_{Lvis} = \frac{-\ln(I_{bnvis}) + \ln(I_{Scvis} E_0)}{\delta_{Rvis} m_r} \quad (35)$$

*3.2.1.15. Cucumo et al.* Cucumo et al. [71] developed a simple and general calculation model for the determination of the average daily Linke turbidity factor in places where the daily global irradiation on the horizontal plane is known. Their model was validated using experimental data for the turbidity factor $T_L$ obtained in Arcavacata di Rende (Cosenza) and in Casaccia (Rome). The model was also experimentally validated using experimental $T_L$ data obtained in Arcavacata di Rende and in Casaccia. Moreover, polynomial correlations of the turbidity factor had been developed for 28 Italian localities where the European Solar Radiation Atlas (ESRA) provided maximum radiation values during clear days. Using these correlations, the average turbidity factor on each day of the year was calculated.

The suggested formula for estimating $T_L$ is given by

$$T_L = \frac{-\ln(I_{hn}) + \ln(I_{Sc} E_0)}{\delta_R m_r} \quad (36)$$

where $I_{hn}$ is the direct hourly normal energy (i.e., the incident hourly energy on a plane perpendicular to the Sun's rays), $\delta_R$ is the result of the formula by Louche et al. [39] corrected by Kasten [69], and $m_r$ is the expression by Louche et al. [39].

*3.2.1.16. Geiger et al.* Geiger et al. [72] summarized the equations that describe the modified European Solar Radiation Atlas (ESRA) model, which was initially presented by Rigollier et al. [73] and then modified by Remund and Page [74] to take into account the variation in the integral optical depth with the altitude of the site.

*3.2.1.17. Ineichen and Perez.* Ineichen and Perez [75] presented a new formulation of turbidity that displays a relatively stable behavior during a specific day. The new turbidity coefficient agreed with that by Kasten [64] at an air mass of 2. However, the contribution of absorption by atmospheric gases into the integral optical depth was not taken into account. The formulation of $T_{LIP}$ can be expressed as follows:

$$T_{LIP} = 1 + \left[11.1 \ln(b E_0 I_{bn})\right]/m_r \quad (37)$$

where $b$ is given by

$$b = 0.664 + 0.163/\exp(\text{station height, meters}/8000) \quad (38)$$

The new formulation has the following advantages: (1) solar altitude-independent and (2) matches the original Linke turbidity factor at the air mass of 2. Therefore, the new formulation remains coherent with previous studies and is based on high quality data that are representative of widely differing geographic locations, altitudes, and climates.

*3.2.1.18. Polo et al.* Polo et al. [32] presented a methodology for estimating daily Linke turbidity factor under clear sky conditions at an air mass of 2 from the global horizontal irradiance data at solar noon and from monthly mean values of the Linke turbidity factor. The method utilized the ESRA clear sky model to establish a functional relationship between the global irradiance at solar noon and the final estimation of the Linke turbidity factor.

### 3.3. Schüepp's turbidity coefficient (B)

Schüepp [76] introduced the turbidity parameter $B$, which is fundamentally based similarly with $\beta$, but referred to the base 10 instead of $e$ and to 0.5 μm wavelength instead of 1 μm.

$$B_{\Delta\lambda} = \frac{\log \frac{1}{E_0} \frac{I_0(\Delta\lambda)}{I(\Delta\lambda)}}{m_r} - (\overline{\tau}_R + \overline{\tau}_w) \quad (39)$$

where $I(\Delta\lambda)$ is the direct solar flux on the Earth's surface, $I_0(\Delta\lambda)$ is the spectral irradiance above the atmosphere, $\overline{\tau}_R$ is the mean

Rayleigh extinction, and $\overline{\tau}_w$ is the selective absorption coefficient for atmospheric gases $O_2$, $O_3$, $H_2O$, $CO_2$, and so on.

#### 3.3.1. Techniques for determining Schüepp's turbidity coefficient (B)

Katz et al. [65] identified relationships and correlations among various atmospheric turbidity coefficients based on the direct radiation measurements in five bands of the visible region, which were performed in a semi-rural site. Tables 2 and 3 present the values of $a$ and $b$ coefficients in the linear regression $Y=a+bX$, which were calculated among different turbidity parameters $\beta$, $B_i$, and $B_{0.5}$.

### 3.4. Unsworth–Monteith's turbidity coefficient ($T_U$)

Unsworth and Monteith [77] introduced a new method for the estimation of the turbidity of the atmosphere. The Unsworth–Monteith turbidity coefficient $T_U$ represents the number of clean and dry atmospheres (CDAs), which exhibit the same attenuation as the actual atmosphere that contains aerosols and water vapor. Typically, $T_U$ varies from 0 to 1. Unsworth and Monteith used experimental irradiance measurements of a clean atmosphere and created the following formulation:

$$I_{bn\lambda} = I^{*}_{bn\lambda}\exp(-T_{U\lambda}m_r) \tag{40}$$

where $I_{bn}$ and $m_r$ are defined earlier, $I^{*}_{bn}$ represents the normal incidence total solar radiation under a dust-free atmosphere (W m$^{-2}$), and $\exp(-T_U m_r)$ represents an aerosol spectral transmittance coefficient. With the definition of the integrated total optical thickness, an integrated expression of $T_U$ that is similar to the expression of $T_L$ can be obtained as follows:

$$T_U = -\frac{1}{m_r}\ln\left(\frac{I_{bn}}{I^{*}_{bn}}\right) \tag{41}$$

### 3.5. Illuminance's turbidity factor ($T_{il}$)

Illuminance turbidity factor $T_{il}$ was introduced by Navvab et al. The concept of the illuminance turbidity factor is analogous to that of the Linke factor with $I_{bn}$, $I_0$, and $T_L$ substituted with $I_{\upsilon sn}$, $I_{\upsilon 0}$, and $T_{il}$. $T_{il}$ is apparently a better parameter for describing atmospheric conditions because, unlike the Linke factor, it is insensitive to water vapor content, air mass, and solar altitude angle or time of day. $T_{il}$ can be determined from the direct measurement of the beam normal illuminance by Chaiwiwatworakul et al. [31] and Navvab et al. [78]. Illuminance turbidity can be obtained as follows:

$$T_{il} = -\left(\frac{1}{\delta_{il}m_r}\right)\ln\left(\frac{I_{\upsilon sn}}{I_{\upsilon 0}}\right) \tag{42}$$

The values of $\delta_{il}$ can be calculated as follows:

$$\delta_{il} = \frac{0.1}{(1+0.0045m_r)} \tag{43}$$

Illuminance turbidity has a heuristic advantage as an approximate analytical expression because it can be directly derived from the Ångström formula as follows:

$$T_{il} = 1+21.6\beta \tag{44}$$

**Table 2**
Values of $a$ in the linear regression $Y=a+bX$.

| Y/X | $\beta$ | $B_1$ | $B_2$ | $B_3$ | $B_4$ | $B_5$ | $B_{0.5}$ |
|---|---|---|---|---|---|---|---|
| $\beta$ | 0 | −0.005 | 0.008 | 0.002 | −0.007 | −0.003 | −0.001 |
| $B_1$ | | 0 | 0.019 | 0.018 | 0.008 | 0.010 | 0.001 |
| $B_2$ | | | 0 | −0.003 | −0.010 | −0.009 | −0.004 |
| $B_3$ | | | | 0 | −0.006 | −0.001 | 0.006 |
| $B_4$ | | | | | 0 | 0.002 | 0.005 |
| $B_5$ | | | | | | 0 | 0.006 |
| $B_{0.5}$ | | | | | | | 0 |

**Table 3**
Values of $b$ in the linear regression $Y=a+bX$.

| Y/X | $\beta$ | $B_1$ | $B_2$ | $B_3$ | $B_4$ | $B_5$ | $B_{0.5}$ |
|---|---|---|---|---|---|---|---|
| $\beta$ | 1 | 0.95 | 1.39 | 0.73 | 0.87 | 0.93 | 1.02 |
| $B_1$ | | 1 | 1.36 | 0.69 | 0.83 | 0.91 | 1.07 |
| $B_2$ | | | 1 | 0.51 | 0.61 | 0.68 | 0.72 |
| $B_3$ | | | | 1 | 1.13 | 1.24 | 1.29 |
| $B_4$ | | | | | 1 | 1.10 | 1.19 |
| $B_5$ | | | | | | 1 | 1.05 |
| $B_{0.5}$ | | | | | | | 1 |

### 3.6. Volz's turbidity factor ($T_V$)

Computations of Volz turbidity are performed according to the usual Bouguer–Lambert–Beer Law that expresses the measured intensity at wavelength $\lambda$ as follows [79]:

$$T_V = -\left[\frac{\ln I_{bn} + \ln E_0 - \ln I_0}{m_r}\right] - \tau^{R}_{\lambda} - \tau^{Oz}_{\lambda} \tag{45}$$

where $\tau^{R}_{\lambda}$ is the Rayleigh scattering coefficient for air molecules at wavelength $\lambda$, and $\tau^{Oz}_{\lambda}$ is the absorption coefficient for ozone at $\lambda$.

### 3.7. Kastrov's turbidity coefficient ($T_K$)

Adeyefa et al. [80] reported that Kastrov [81] obtained the following simple formula for the integral solar radiation on the surface level:

$$I_b = \frac{I_{Sc}}{(1+T_K m^*)} \tag{46}$$

from which the following can be obtained:

$$T_K = \frac{1}{m} \times \left(\frac{I_{Sc}}{I_b} - 1\right) \tag{47}$$

where $T_C$ is a quantitative characteristic of atmospheric transparency.

### 3.8. Rational's turbidity factor ($T_r$)

Majumda et al. [82] defined a new measure of total atmospheric turbidity, termed as the rational turbidity factor, $T_r$, to overcome the limitations of the Linke turbidity factor $T_L$. $T_r$ is composed of three components, namely, (i) pare and dry air (the basic effect), (ii) precipitable water vapor $w$, and (iii) aerosol particles in the atmosphere (dust, smoke, and haze). The formulation of $T_r$ is expressed as follows:

$$T_r = \frac{1}{m_r}\left(\frac{0.32491 - \log I_{bn}}{0.072375}\right)^{1/0.57} \tag{48}$$

Majumda et al. [82] correlated the rational turbidity factor $T_r$ with the Ångström–Schüepp turbidity coefficient $B$ for different values of precipitable water $w$ at unit air mass to eliminate the effect of any possible virtual variation with air mass. They demonstrated that the effect of $B$ and $w$ were inseparable because the scattering of radiation by aerosol particles and the absorption of the same by water vapor coexist within the absorption bands of water vapor. Thus, the two effects cannot be isolated from each other. This interaction becomes more pronounced with increasing

particle size since the negative exponent of $\lambda$ in the Ångström–Schüepp equation approaches zero for large particles.

Majumda et al. [83] also investigated the effects of varying air masses $m_r$ on $T_r$ in relation to $B$ and $w$ and presented a formula that connects $T_r$ with $B$, $w$, and $m_r$ as follows:

$$T_r = 1 + 80(B/1.8)^{1+0.817(0.586)m_r} + \left[25\left(\frac{B+0.1}{1.9}\right)^{1-0.2347(0.869)m_r}\right] w^{0.3} \quad (49)$$

## 4. Conclusion

The results of this review indicate the following:

1. In the last century, a number of atmospheric turbidity indices were introduced and several techniques were developed to determine their values. The most currently used indices include the Ångström turbidity coefficient $\beta$, the Linke turbidity factor $T_L$, the Schüepp turbidity coefficient $B$, and the Unsworth–Monteith $T_U$.
2. The Ångström exponent $\alpha$ is very popular because the simplicity of its equation enables the interpolation or extrapolation of aerosol optical properties and because of its connection to particle microphysics. In addition, the Ångström exponent $\alpha$ also indicates the amount of aerosols and represents the combined effects of scattering and absorption caused by aerosols.
3. The Ångström turbidity coefficient $\beta$ has long been used as a turbidity index and it can be determined by different methods for spectral and broadband radiation measurements in large spectrum bands that are not affected by water vapor absorption (280 nm $< \lambda >$ 630 nm).
4. $\beta$ cannot be evaluated unless $\alpha$ is known a priori. This uncertainty in $\alpha$ makes the determination of $\beta$ less precise and discriminating.
5. The Linke turbidity factor $T_L$ is important for the calculation of the beam solar and the diffused radiation in a specific locality. This parameter is also useful for the prediction of available solar radiation and daylight illuminance under cloudless skies. The Linke turbidity factor $T_L$ is also a key input to several models that assess down-welling irradiance under clear skies and that are used by several communities in the fields of renewable energies, climatology, agro-meteorology, and atmospheric pollution.
6. The main problem with $T_L$ is that it is difficult to be accurately determined and it fails to distinguish the effect of water vapor from the absorption caused by aerosol. $T_L$ is also not truly independent of the optical air mass and it varies diurnally even under an unchanging atmosphere. Moreover, $T_L$ is subject to virtual daily variations, such as broadband aerosol optical depth $\delta_a$ caused by the combination of different parasitic effects. Virtual variations in $T_L$ caused by the solar zenith angle $\gamma$ are smoother compared with those of $\delta_a$ for $\gamma < 80^\circ$, but become steeper for larger zenith angles. A reference air mass value of 2 should be calculated to remove a part of the $T_L$'s parasitic variation.
7. The Ångström and the Schüepp turbidity coefficients, $\beta$ and $B$ respectively, have the same spectral definition. They both correspond to $\tau_{\alpha\lambda}$ at 1000 nm (base $e$) and at 500 nm (base 10), respectively, and thus, they are *true* turbidity coefficients that are unaffected by anything except by the aerosol total burden.
8. The Unsworth–Monteith coefficient (or broadband aerosol optical depth) was found to be slightly dependent on both zenith angle and water vapor. Hence, caution should be observed when comparing data from different parts of a day or year.

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